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Mantle flow influence on subduction evolution

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ABSTRACT

The impact of remotely forced mantle flow on regional subduction evolution is largely unexplored. Here we investigate this by means of 3D thermo-mechanical numerical modeling using a regional modeling domain. We start with simplified models consisting of a 600 km (or 1400 km) wide subducting plate surrounded by other plates. Mantle inflow of ∼3 cm/yr is prescribed during 25 Myr of slab evolution on a subset of the domain boundaries while the other side boundaries are open. Our experiments show that the influence of imposed mantle flow on subduction evolution is the least for trench-perpendicular mantle inflow from either the back or front of the slab leading to 10–50 km changes in slab morphology and trench position while no strong slab dip changes were observed, as compared to a reference model with no imposed mantle inflow. In experiments with trench-oblique mantle inflow we notice larger effects of slab bending and slab translation of the order of 100–200 km. Lastly, we investigate how subduction in the western Mediterranean region is influenced by remotely excited mantle flow that is computed by back-advection of a temperature and density model scaled from a global seismic tomography model. After 35 Myr of subduction evolution we find 10–50 km changes in slab position and slab morphology and a slight change in overall slab tilt. Our study shows that remotely forced mantle flow leads to secondary effects on slab evolution as compared to slab buoyancy and plate motion. Still these secondary effects occur on scales, 10–50 km, typical for the large-scale deformation of the overlying crust and thus may still be of large importance for understanding geological evolution.

1. Introduction

The dynamical and morphological evolution of subducting slabs has for decades been widely investigated by means of 2D and eventually 3D numerical and laboratory modeling. 3D modeling brought focus on the feedback between slab-induced mantle flow and slab evolution with particular attention for the trench-parallel flow and toroidal flow occurring around the edges of a retreating slab (e.g. Kincaid and Griffiths, 2003; Funiciello et al., 2003, 2004, 2006; Schellart, 2004; Piromallo et al., 2006; Stegman et al., 2006; Schellart et al., 2007; Stegman et al., 2010; Butterworth et al., 2012; Capitanio and Faccenda, 2012; Jadamec and Billen, 2012; Schellart and Moresi, 2013; MacDougall et al., 2014; Crameri and Tackley, 2014; Sternai et al., 2014). These 3D modeling studies have demonstrated the importance of mantle flow that is induced by slab motion, but how remotely forced mantle flow affects slab evolution remains unexplored. By remotely forced, or external, mantle flow we mean flow that is caused by plate motions and subduction elsewhere or is excited by deeper mantle processes such as rising plumes or sinking of detached slab. Slab-induced flow resulting from e.g. slab rollback can be perceived as superposed on, or interacting with this background flow. Until now only a few papers (Hager et al., 1983; Olbertz et al., 1997; Boutelier and Cruden, 2008; Winder and Peacock, 2001; Rodríguez-González et al., 2014; Ficini et al., 2017) paid attention to the possible influence of remotely forced mantle flow on subduction evolution. These studies demonstrate a strong influence on slab geometry evolution but due to the 2D nature of the modeling, not allowing for toroidal mantle flow, such inferences are restricted to geodynamic settings in which poloidal flow strongly dominates.

Here we investigate the impact of remotely forced mantle flow on subduction evolution with 3D thermo-mechanical numerical modeling using a regional modeling domain by imposing mantle inflow on a subset of the side boundaries. As this topic is unexplored in 3-D our primary purpose is to browse through the model space searching for first-order effects. We monitor the evolution of slab morphology under different external mantle flow conditions both for generic subduction models and for a simulation of natural subduction in the western Mediterranean. For

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the generic simulations, we use simplified models of subduction evolution comprised of rectangular plates and a straight trench with initial trench-perpendicular plate convergence. This is done to make a link with this often-used model setup in 3-D subduction modeling. We include in our experiments the overriding plate and plates adjacent to the subducting plate, similar to recently used model setups (e.g. Yamato et al., 2009; Capitanio et al., 2010; Butterworth et al., 2012; Meyer and Schellart, 2013; Boutelier and Cruden, 2013; Guillaume et al., 2013; Moresi et al., 2014). We combine side boundaries with prescribed mantle flow with side boundaries that are open (Chertova et al., 2012, 2014a, 2014b) for conservation of mass. Because free-slip boundary conditions are most often used in 3D subduction modeling using a regional model domain, we briefly address the magnitude of the difference between using either boundary condition in the experiments we conduct here. Lastly, we investigate the influence of remotely forced mantle flow in a simulation of the Rif–Gibraltar–Betic subduction evolution since ~35 Ma in the western Mediterranean region. This builds on our earlier work for this subduction system (Chertova et al., 2014a, 2014b) in which we used open side boundaries. In these experiments, we impose time-dependent boundary conditions of mantle flow determined from back-advection of the present-day temperature and density structure of the mantle derived from a tomographic model. These experiments also serve to test the robustness of the earlier modeled subduction evolution (Chertova et al., 2014a, 2014b) with respect to the viscous forcing exerted on the slab by remotely forced mantle flow.

2. Methodology

We carry out our 3-D numerical experiments using the finite element package SEPRAN (Segal and Praagman, 2005). As described in Chertova et al. (2014a, 2014b), we solve the equations of mass, momentum and energy conservation and the transport equation for advection of non-diffusive material properties using the extended Boussinesq approximation including two major phase transitions at 410 km and at 660 km in the unperturbed mantle. Experiments are performed in a Cartesian box with dimensions of length × width × height = 3000 km × 2000 km × 1000 km for the generic model setup and 1800 km × 1300 km × 1000 km for the simulation of western Mediterranean subduction.

2.1. Rheology and boundary conditions

Composite nonlinear rheology is used for all models and comprises diffusion creep, dislocation creep, and a viscosity maximum, $\eta_{\max}(2\times10^{23}$ or 5×10^{23} Pas). For the geometrically more complex western Mediterranean model setup also a stress limiter mechanism is used (Chertova et al., 2014a, 2014b). The effective composite viscosity is defined as $\frac{1}{\eta_{eff}} = \frac{1}{\eta_{diff}} + \frac{1}{\eta_{disl}} + \frac{1}{\eta_y}$, with

$$\eta_{diff} = \mu A_{diff}^{-1}(b/d)^{-m}\exp\left[(E_{diff}+PV_{diff})/RT\right] \tag{2.1}$$

$$\eta_{disl} = \mu A_{disl}^{-\frac{1}{n}}\dot{\varepsilon}^{\frac{1-n}{n}}\exp\left[(E_{disl}+PV_{disl})/nRT\right] \tag{2.2}$$

$$\eta_y = \frac{\tau_{\max}}{2\dot{\varepsilon}}, \tag{2.3}$$

where μ is the shear modulus, $A_{diff,disl}$ are diffusion and dislocation creep viscosity prefactors, b is Burgers vector, d is the grain size, $\hat{\varepsilon}$ is the second invariant of the strain-rate tensor, m is the grain size exponent, $V_{diff,disl}$ and $E_{diff,disl}$ are activation volume and activation energy for diffusion and dislocation creep, respectively, P is the lithostatic pressure, T is temperature, $\tau_{\max}$ is the maximum yield stress value, i.e. stress limiter. Used parameters for the rheology and phase transitions are given in supplementary Table 1. Lagrangian particles are used to define viscosity parameters. They are placed randomly over the whole modeling domain and free inflow/outflow of particles is allowed on side boundaries. The number of particles varies between 20 and 30 million during computations. We use free-slip conditions for the top boundary and no-slip for the bottom boundary. Different side boundary conditions are used: free-slip (closed, impermeable boundary), open boundaries as developed in Chertova et al. (2012), or prescribed mantle flow at a subset of the vertical sides, as detailed in the next section. For all boundaries we prescribe a stationary depth-dependent adiabatic temperature profile during simulations. The mesh element size varies from 6 km in the top 200 km to 20 km at the bottom of the model.

To allow decoupling of the subducting slab from the free-slip top surface and overriding plate we implement a 30 km weak crustal layer, as in Chertova et al. (2012, 2014a, 2014b), with a viscosity 2×10^{19} Pas in the generic models and 5×10^{19} Pas for the western Mediterranean subduction modeling. This weak top layer facilitates vertical decoupling of the slab during the development of subduction, which is driven by its slab buoyancy, and by the subducting plate motion that is imposed on the side-boundary, both simulating geodynamic forcing that also occurs naturally. Additional forcing comes from the mantle flow imposed on selected boundaries of the model of which the effects on subduction evolution are topic of our investigation here.

2.2. Initial model setup for generic models

The initial model setup is illustrated in Fig. 1A and comprises the subducting and overriding plates and two plates to either side. Table 1 lists the models we created with the various inflow boundary conditions. All 4 plates have the same initial thermal structure determined from the equation of a cooling semi-infinite half space for an oceanic lithosphere age of 80 My. The initial subduction angle is 45° and the slab extends to the depth of 300 km. Low viscosity deformation zones of $3\cdot10^{20}$ Pas and 70 km wide (pink in Fig. 1A) are used as a weak mechanical coupling of the plates (Fig. 1B). Fig. 1B shows their motion decoupling effect between subducting and side-plates in the initial model stage. Experiments are performed with a subducting plate width of 600 and 1400 km. These models include a viscosity increase in the mantle by a factor of ~10 to 10^{22} Pas at a depth of ~660 km. The upper viscosity limit is set to $5\cdot10^{23}$ Pas. The activation energy for dislocation creep and activation volume for diffusion creep for most models is set to 423 $\mathrm{KJ\,mol^{-1}}$ and 3 $\mathrm{cm^3\,mol^{-1}}$ respectively. But, for models M600.O*, M600.SE, M600.SW and M600.O/E/S a slightly higher activation energy of 433 $\mathrm{KJ\,mol^{-1}}$ is used for reasons explained later. An example of viscosity profiles away from the subducting slab is shown in supplementary Fig. 1. In all models, we prescribe the speed of the subducting plate of 1.5 cm/yr on the western side boundary, which ensures that the slab does not rollback too fast to this side boundary. Although this choice is practical, it is not unnatural because slab pull and an independent component of advance velocity of the subducting plate are natural drivers of many subduction zones (e.g. Heuret and Lallemand, 2005; Schellart, 2008).

For model comparison, we constructed two reference models of subduction evolution, M600.O and M1400.O, in which 4 open boundaries are used and thus no mantle inflow is imposed. In separate experiments uniform inflow of the mantle at a speed of 3 cm/yr is prescribed below the lithosphere on either the western (left), eastern (right) or southern (frontal) model boundary keeping other boundaries open (Table 1). Transitions between boundary conditions are taken up locally and smoothly by the modeled mantle flow. Lastly, we created experiments where the prescribed mantle flow of 3 cm/yr gradually changes direction from eastward